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The Tibetan Plateau cryosphere: Observations and model simulations for current status and recent changes

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ABSTRACT

Global warming has already had a significant impact on social ecosystems. The Tibetan Plateau (TP), which is characterized by a cryosphere, is also recognized to have a profound influence on regional and global climate systems, as well as the ecological economy. Therefore, research on the cryosphere is of significant importance. This paper comprehensively reviews the current status and recent changes of the cryosphere (e.g., glacier, snow cover, and frozen ground) in the TP from the perspectives of observations and simulations. Because of enhanced climate warming in the TP, a large portion of glaciers have experienced significant retreat since the 1960s, with obvious regional differences. The retreat is the smallest in the TP interior, and gradually increases towards the edges. Glacier simulations are comparatively few and still under development. Snow cover is a highly sensitive element of the cryosphere and decreases with large interdecadal variations from the 1960s to the 2010s in general. Simulations of snow cover mostly focus on the mutual feedback between the snow cover anomaly, climate and atmospheric circulation. In situ observations and simulations both indicate that the mean annual temperature of frozen ground increases, causing permafrost thaw and degradation and decreasing the seasonal freeze depth of seasonally frozen ground. Under future climate warming, the cryospheric elements in the TP will continue to diminish on the whole. Studies of climate and the cryosphere are ongoing. To date, the lack of observations is the biggest challenge on the TP, resulting in a divergence of cryosphere dynamics and its simulation being a bottleneck. To overcome these issues, a strategy that combines sets of in situ and remote sensing measurements and improved numerical models is of great importance for achieving breakthroughs with respect to research on the TP cryosphere and its interaction with climate.

1. Introduction

The cryosphere, which is a special circle, is comprised of solid water (snow, river and lake ice, sea ice, glaciers, ice caps, ice shelves, ice sheets, and frozen ground) under a certain low temperature on earth (Stocker et al., 2013). It constitutes five layers on the crustal surface together with the lithosphere, atmosphere, hydrosphere, and biosphere. At present, 7% of the global ocean area and 11% of the land area is covered by multiyear ice. Up to 24% of Northern Hemisphere land area is occupied by permafrost (Zhang et al., 2000). The extent of seasonal snow and ice accounts for 15% of the land area in January, and 9% in July. Seasonally frozen ground (SFG) is distributed more widely (Shi and Cheng, 1991). With a unique position in global change, the cryosphere is a component of the climate system and is also a product of climate. Once it comes into existence, the cryosphere exerts huge effects and feedback on climate by influencing surface energy, water vapor flux, clouds, precipitation, hydrology, the atmosphere, ocean currents, and carbon cycle processes (Schmittner et al., 2002; Schuur et al., 2009; Immerzeel et al., 2010; Flanner et al., 2011; Mölg et al., 2014). The temperature increase in cryospheric areas (high mountain regions) is greater than in other areas of the globe (Pepin et al., 2015). The cryosphere is, therefore, the most sensitive to climate change. One of the most powerful pieces of evidence for global warming is the rapid variation of the cryosphere during recent decades. A positive feedback mechanism in the cryosphere enhances climate warming (Li et al., 2008; Xiao et al., 2008; Stocker et al., 2013).

Because of its functions and profound influence, the cryosphere has gained extensive attention from international research projects and

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academic organizations. During the last several decades, the International Commission on Snow and Ice (ICSI) has made unremitting efforts for Cryospheric Science to get a better position in the International Union of Geodesy and Geophysics (IUGG). The World Climate Research Programme (WCRP) established a subproject called Climate and Cryosphere (CliC) in March 2000, and it started in 2001 (Allison et al., 2001). The aim of the programme includes the following: improving understanding of physical process and feedback mechanisms between the cryosphere and climate systems; increasing the accuracy of models describing cryospheric processes; reducing the uncertainty in climate simulation and prediction; and assessing and quantifying the changes of each component in the cryosphere caused by past and future climate changes. In 2004, the WCRP, CliC project and the International Council for Science cosponsored the Integrated Global Observing Strategy (IGOS), and cryosphere monitoring was one of the main observation projects. In 2007, the IUGG established the International Association of Cryospheric Sciences (IASC) to promote the research of physical processes in the global cryosphere (Jones, 2008). Under the background of international programs in terms of the cryosphere and environment, a series of major plans such as the Third Pole Environment (TPE) plan (Yao et al., 2012a) and the first and second Comprehensive Scientific Expeditions to the Tibetan Plateau (TP) have been initiated by Chinese scientists in recent years to investigate cryosphere changes and their influence on regional climate, water resources, and socially sustainable development.

The cryosphere is mainly distributed at high latitude in polar regions, but it is also widespread on plateaus and high mountains at middle and low latitudes. China hosts the largest cryosphere concentration at middle and low latitudes. The second Chinese Glacier Inventory reported that China currently has 48,571 glaciers, covering a total area of $5.18 \times 10^4\,km^2$, occupying 7.1% of the global glacial area outside the Antarctic and Greenland (Liu et al., 2015). The permafrost area is estimated at $\sim 1.59 \times 10^6\,km^2$, the SFG (excluding instantaneous frozen ground) is $\sim 5.36 \times 10^6\,km^2$ (Ran et al., 2012), and the extent of stable snow cover (snow cover over 60 d) is approximately $4.20 \times 10^6\,km^2$ (Qin et al., 2006a). The cryosphere in China mainly contains glaciers, snow cover, and frozen ground, and it is extensively distributed in the TP (Xiao et al., 2008; Yang et al., 2010; Yao et al., 2012b).

With an average height of over 4000 m, the TP is known as the world's 'Third Pole'. It stretches from the Pamirs to the Hengduan Mountains and spans 31° of longitude (26°00′12″N–39°46′50″N), or approximately 2945 km from west to east. It stretches from south of the Himalayas to north of the Kunlun-Qilian Mountains and spans 13° of latitude (73°18′52″E–104°46′59″E), or approximately 1532 km from south to north. The area of the TP is approximately $2572.4 \times 10^3\,km^2$, accounting for 26.8% of the total land area of China (Zhang et al., 2002) (Fig. 1a). The TP imparts significant impact on Asian monsoon circulation, precipitation in the middle and lower reaches of the Yangtze River in China, and even on the global climate through thermal and mechanical dynamics (Wu et al., 2012a). The TP is also called the "water tower of Asia", because it brings glacial melt water to rivers, which is particularly crucial for water resources on the Asian continent affecting the livelihoods of > 1.4 billion people (Immerzeel et al., 2010). Widespread cryosphere in the TP (glacier, snow cover, permafrost, and SFG) is highly vulnerable to climate forcing and has vital feedback to climate change. In the context of global warming, the cryosphere has indeed undergone considerable change and received much attention recently. In this paper, we provide an overview of recent studies on the TP cryosphere. The article structure is organized as: climate change focusing on temperature and precipitation on the TP during recent decades is shown in Section 2; the cryosphere (glacier, snow cover, and frozen ground) in the TP is reviewed in Section 3, of which each cryosphere element is presented starting from its current monitoring, distribution, and recent changes from observations and model simulations perspectives. Section 4 discusses the climate-cryosphere interactions in the TP as far as we are concerned. Lastly, a summary and possible prospects are given in Section 5.

2. Climate change on the TP

The cryosphere is a sensitive and visual indicator of climate change. The advance and retreat of glaciers, the increase and decrease of snow cover and frozen ground are all largely related to temperature and precipitation. Therefore, research on the TP climate is extremely imperative.

During the last century, especially since the 1960s, global warming as the main feature of climate change on the TP has become increasingly significant (e.g., Liu and Chen, 2000; Wang et al., 2008; Xu et al., 2008; Guo and Wang, 2011; Chen et al., 2013; Yang et al., 2014; Kuang and Jiao, 2016; Pang et al., 2017). Although the study time period and selected meteorological stations are different, the annual average air temperature on the TP presents an increasing trend with varying degrees. From 1960 to 2010, the rate of temperature rise was 0.20 °C $decade^{-1}$ (Chen et al., 2013, see Fig. 2). Compared with the Northern Hemisphere and the global averages, the warming of the TP is earlier and greater (Liu and Chen, 2000). The temperature increase reached 0.3–0.4 °C $decade^{-1}$ in the past 50 years (1960–2012), which is more than twice the global temperature rise in the same period (Chen et al., 2015). In winter, the rate of temperature increase is higher (Liu and Chen, 2000; Wang et al., 2014a) and it is almost double the rate in spring (Lu and Liu, 2010). The maximum and minimum air temperatures exhibit rising trends and the rate of rising minimum temperature is more than twice that of the maximum temperature (Liu et al., 2006a). The warming in the central, eastern, and northwestern TP is obvious (Liu and Chen, 2000). Some studies showed a pronounced warming in the northern region, with a rate higher than other regions (Li et al., 2010a; Guo and Wang, 2011; Wang et al., 2014a, 2017b). The number of extremely cold days, cold nights, frost days, and frozen days on the plateau decreased significantly, the number of extremely warm days and warm nights and the length of growing season increased significantly, and the diurnal temperature range decreased significantly (You et al., 2008). The most significant decrease is the frequency of extreme low temperature events in the northern plateau (Li et al., 2010a).

Overall, a large-scale climate warming trend appeared on the TP during the last half century that began earlier and is greater than those of the Northern Hemisphere and the global averages. The climate warming has obvious regional differences with the most remarkable warming in the northern region. The increasing temperature occurs across all four seasons and is especially noticeable in winter. Moreover, the maximum and minimum air temperatures both have a tendency to increase. The minimum temperature increase rate is higher than for the maximum, leading to a decreasing diurnal temperature range on the TP. As a consequence, extremely cold events decrease and warm events increase.

Compared with temperature, precipitation change over the TP is more complex (Yang et al., 2007a). The main spatial pattern of summer precipitation is a seesaw structure from north to south. Obvious differences in precipitation exist between the north and south plateau. In particular, between the south and the northeast where precipitation changes are even inverse (Liu and Yin, 2001; Yang et al., 2007b). The trend of precipitation has been positive in most regions during the past decades (Xu et al., 2008; Gao et al., 2015a; Wang et al., 2018a), i.e., in the central and northeastern TP covering the Qaidam Basin (Wang et al., 2014a) and Qilian Mountains (Wang et al., 2018b). However, the trend in the western, eastern, southern, and southeastern parts of the TP is negative (Li et al., 2010a; Chen et al., 2013; Yang et al., 2014; Gao et al., 2015a, Fig. 2). The increase rate of precipitation for the whole TP was 3.11 mm $decade^{-1}$ ($P = .1936$) from 1960 to 2010 (Chen et al., 2013). The gridded precipitation data revealed that the average annual precipitation increased with fluctuations from 1961 to 2012 at a rate of

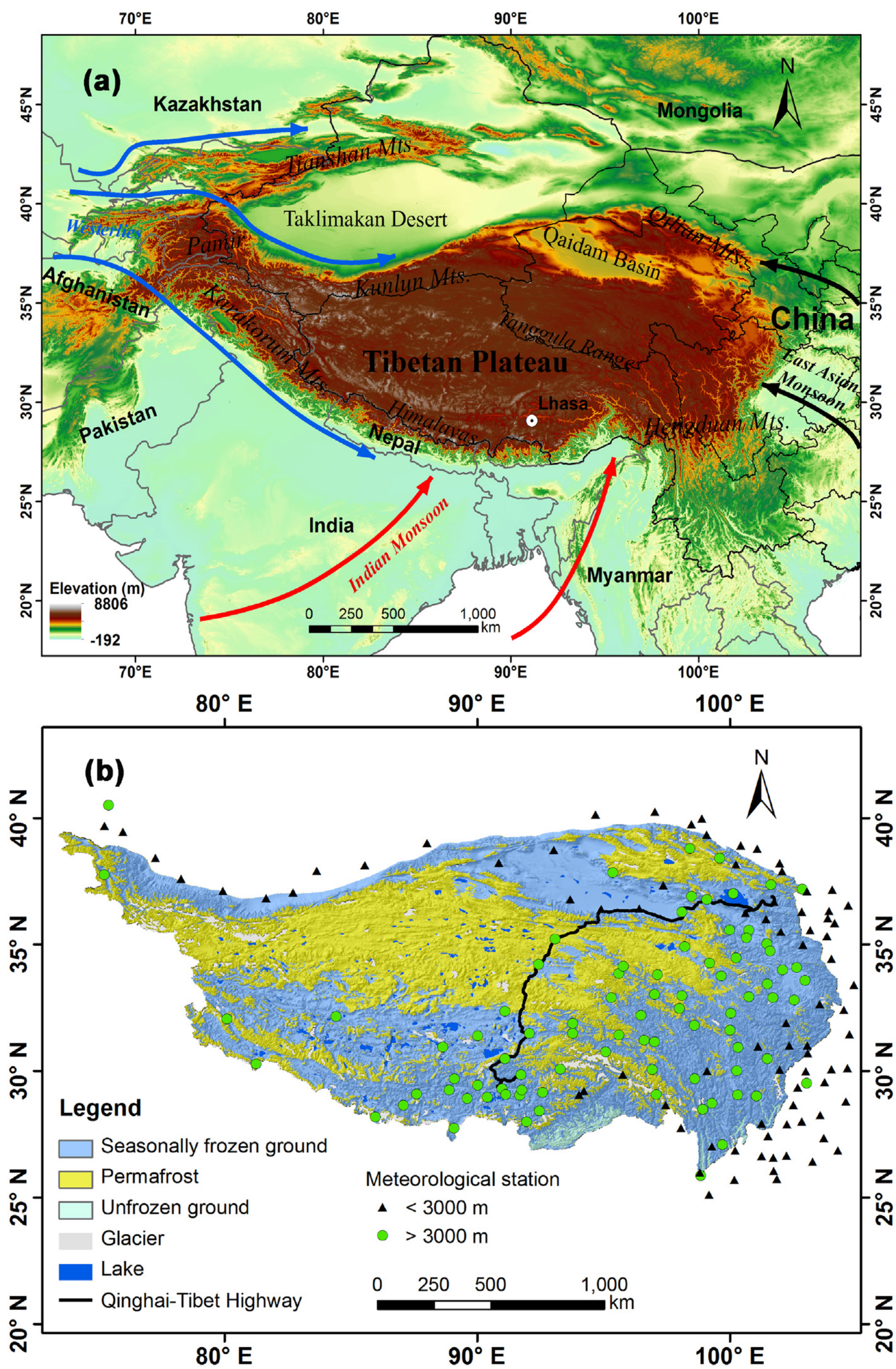

Fig. 1. (a) Topographic map and dominant atmospheric circulations (the Indian monsoon, East Asian monsoon, and westerlies) affecting the TP; (b) distribution of frozen ground, glaciers (after Zou et al., 2017), and meteorological stations in the TP.

5.07 mm decade^{-1} (Wang et al., 2018a). On a seasonal basis, for the entire TP, the most significant increasing trend appears in spring. Winter precipitation is also shown to significantly increase (Fig. 2b), whereas summer and autumn have non-significant downward precipitation trends (Chen et al., 2013), contrary to recent relevant studies demonstrating slight upward trends (Gao et al., 2015a; Kuang and Jiao, 2016; Wang et al., 2018a), mainly owing to the difference in data selection and the uncertainty of precipitation study. As the main form of precipitation on the TP, snowfall accounts for approximately 24% on multiyear timescales over frozen ground regions and has an even higher proportion during winter and spring (Zhu et al., 2017). Snowfall increased on the central and western TP but deceased on southeastern and northeastern parts during 1960–2014. Region-wide snowfall tended to increase before the 1980s but afterwards began to rapidly decline over most of the plateau, especially on the eastern and northeastern areas (Deng et al., 2017). Along high mountain ranges in the northern TP, snowfall decreased during the wet and warm season from 1957 to 2009 (Cuo et al., 2013). The high-altitude regions are generally cold and increasing temperature can bring more snow. However, where it is already near the freezing point, increasing temperature transforms snow to rain, producing a rapid decline in snowfall (Deng et al., 2017). The frequency of extremely dry events is reduced on the southwestern TP and west of Sichuan province (Li et al., 2010a). In summary, on the whole, the TP precipitation has experienced an increasing trend since

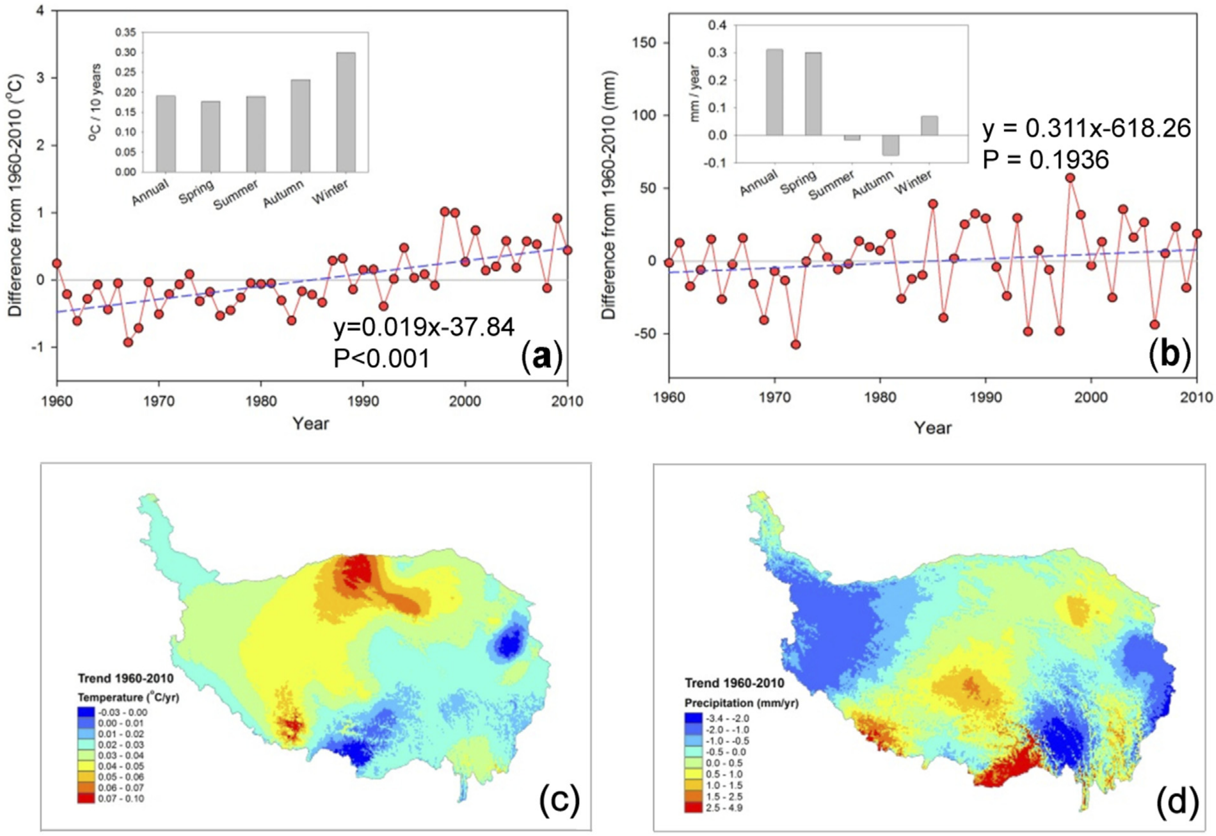

Fig. 2. Observed climate change on the TP from 1960 to 2010. Anomaly time-series for mean annual temperature (a) and precipitation (b). The inset shows trends in seasonal temperature (°C decade^{-1}) (all significant at the p = .01 confidence level) and precipitation (mm yr^{-1}) (spring: p = .01, summer: p = .94, autumn: p = .49, winter: p = .01). Spatial trends in temperature (c) and precipitation (d) over the TP during 1960–2010 (Chen et al., 2013).

the 1960s but with varied patterns. Spatially, an increasing trend occurs over most parts of the TP and a decreasing trend occurs on the rest. The precipitation change exhibits obvious seasonality with a clearly significant positive trend in spring and winter.

Under climate warming, the TP precipitation is expected to increase, following the law that atmospheric water vapor holding capacity rises with temperature (Trenberth, 2011). Different atmospheric circulation patterns are probably the driving forces for regional differences of the TP precipitation change (Yao et al., 2012b). The precipitation at climate scale is mutually and competitively affected by the East Asian and South Asian Monsoons and the mid-latitude westerlies (Schiemann et al., 2009; Feng and Zhou, 2012; Fig. 1a). The negative trend of precipitation on the southern and southeastern TP is likely related to the weakening of the Asian summer monsoon (Yang et al., 2014; Roxy et al., 2015). Based on the comprehensive review of Wang et al. (2018a) and references therein, the strength and nature of the tropical oceanic processes such as the El Niño/Southern Oscillation (ENSO) and Indian Ocean Dipole (IOD) mode along with mid-high latitude atmospheric circulation such as the North Atlantic Oscillation (NAO) impose broad impacts on atmospheric circulation patterns and subsequently modulate water vapor transport to the TP. However, the hinterland of the TP is less affected by the Indian monsoon and westerlies and more controlled by associated local moisture recycling, such as the variation of evapotranspiration induced by vegetation growth (Shen et al., 2015) and energy-water cycles caused by cryospheric processes (snow cover and soil freezing/thawing processes) discussed below.

3. Cryosphere in the TP

3.1. Glaciers

3.1.1. Glacier monitoring

The TP is the region with the most glaciers developed in mid-and low-latitudes. Mountains on the edge of the TP, such as the Himalayas, the Karakorum-Kunlun, and the Kunlun Mountains, possess the most concentrated and the largest number of glaciers. The total area and scale of the glaciers in the central TP is not as large as in the edge mountains. In the Qiangtang Plateau, high mountains above the plateau surface like the Purog Kangri have developed the largest ice cap in mid- and low-latitude regions. Others such as the Zangsegangri, the Buruogangri, and the Malan Mountains have developed many large ice cap type glaciers (Pu et al., 2004; Brun et al., 2017). In 1958, the Chinese Academy of Sciences established a specialized scientific research institution called the Lanzhou Institute of Glaciology and Geocryology (now called Cold and Arid Regions of Environmental and Engineering Research Institute, CAREERI) to research glaciers in mountain regions. After that, several glacier field observation sites, such as the Tianshann, Yulong, and Qilian Mountain stations established by the CAREERI, as well as the Nam Co, Qomolangma, Southeast Tibet, Muztag Ata, and Ngari stations established by the Institute of Tibetan Plateau Research, were successively built to monitor glacier dynamics under the support of the Chinese Academy of Sciences and with the joint efforts of scientists.

The studies of contemporary glaciers are derived from various methods, including field investigation, statistical analysis of observations, experiments and analysis, and records in documents and literature. In the TP, few glaciers can be observed continuously and systematically. The changes of most glaciers are discovered through the

total changes in a certain period or two obtained from the comparison between aerial photos, topographic maps, and records in the literature and field trips. The aerial photos photographed in the 1950s to the 1980s, published aerial topographic maps and professional glacier maps are the critical basic data for studying recent glacier changes (Pu et al., 2004). Other ground-based observations are distributed sporadically in glacier covered areas in the western TP (Xiao et al., 2007). Nevertheless, because of the high altitude and harsh climate conditions, and low density of ground transport networks and logistical support, the surface observations of some regions in the TP are tough to complete. Satellite remote sensing (e.g., Landsat MSS/TM/ETM+, Hexagon KH-9, GLAS/ICESat, and SRTM3 DEM) conjoined with GIS technology have become practical tools for analyzing the status and ongoing glacial change (Liu et al., 2006b; Wang et al., 2013c; Brun et al., 2017). Satellite remote sensing is highlighted to monitor glacier terminal position, area, and volume. For contemporary glaciers in China, the statistical estimation is obtained from topographic maps, Landsat TM/ETM+, and global ASTER photos (Liu et al., 2015). As a result of the westerlies and Indian monsoon, a high concentration and low equilibrium-line altitude (ELA) for glaciers was developed in the southeastern TP and the eastern Pamirs. In central regions of the Pamirs, limited water vapor led to a distribution of high glacier ELAs (Yao et al., 2012b). The number of explored glaciers in the TP within Chinese territory is 46,298, the total glacier area is approximately 59,406 km^2, and the total glacier volume approximately 5590 km^3 (Yao et al., 2004). The total glacier area of the TP and surroundings is about 100,000 km^2 (Yao et al., 2012b). The area and volume of each glacier can be seen in Fig. 3.

3.1.2. Research on recent glacier changes

Mountain glaciers are key indicators of climate change. Over the last hundred years, although there were two stages when the rate of deglaciation decreased or was relatively stable, and some glaciers even slightly advanced (1920s–1930s; 1970s–1980s) (Pu et al., 2004), the general process still presents a clear retreating tendency. Following climate warming, especially since the 1980s, the glacier terminus retreat has accelerated. The largest variability occurs in the glacier terminal positions in the eastern and southern TP, whereas relatively stable positions are located in the northern-central TP and Qiangtang regions, and a high sensitivity of glaciers to climate change occurs in the plateau margin (Pu et al., 2004; Yao et al., 2004). Yao et al. (2004) suggested that the area of glaciers has decreased by 3790 km^2 over the last 40 (1960s–2000) years, meaning that the average glacier thickness thins at an annual rate of 0.2 m. The monsoonal glaciers with temperate characteristics are shrinking. For instance, the total area of the monsoonal temperate glaciers has shrunk by 3921.2 km^2 since the Little Ice Age (Su and Shi, 2000). From 1966 to 2009, the area of 74 monsoonal temperate glaciers in the southeastern TP has decreased by 11.3%, the area and length have separately decreased by 0.8 km^2 and 1146.4 m for the Hailuogou glacier, 2.1 km^2 and 501.8 m for the Mozigou glacier, and 2.4 km^2 and 1002.3 m for the Dagongba glacier (Pan et al., 2012). The stable isotope δ^{18}O in the 41.6-m ice core drilled at 7010 m from the Muztag Ata in eastern Pamir showed a clear tendency of warming in the 1990s. The fast warming in this region at high altitude has already led to the high-speed retreat of mountain glaciers (Tian et al., 2006). Glacial response to plateau climate change is a complicated pattern. For example, the geodesic survey showed that the surface mass balance was -0.31 ± 0.08 m yr^{-1} (1970–2007) in the Mt. Everest region (Bolch et al., 2012) and $-0.7\sim -0.85$ m yr^{-1} (1992–2004) in the Himalayas (Berthier et al., 2007). The Karakoram glaciers did not decline in the early 21st century and had a slight mass gain with a rate of 0.11 ± 0.22 m yr^{-1} (1999–2008) (Gardelle et al., 2012). The area of glaciers in the Yangtze River source region has decreased by 45.75 km^2 from 1964 to 2010 with a relative change of 6.80% (Wang et al., 2013b). An integrated assessment of glacier status in and around the TP in the past 30 years with regard to glacier retreat for 82 mountain glaciers, area reduction for 7090 glaciers, and mass balance for 15 glaciers demonstrated that the mode of mass balance in glaciers is spatially heterogeneous with a net mass loss (Yao et al., 2012b). In the Himalayas (excluding the Karakoram Mountains), glacier retreat is the strongest marked by the largest reduction in glacial length and area and the most obvious negative mass balance. In general, the retreat decreases from the Himalayas to the interior of the TP. The least retreat occurs east of the Pamir, manifesting the smallest length and area reduction and the slightest negative mass balance (Yao et al., 2012b; Fig. 4). Actually, the region-wide mass balance was 1.4 ± 0.8 Gt yr^{-1} in the Kunlun mountains between 2000 and 2016 (Brun et al., 2017).

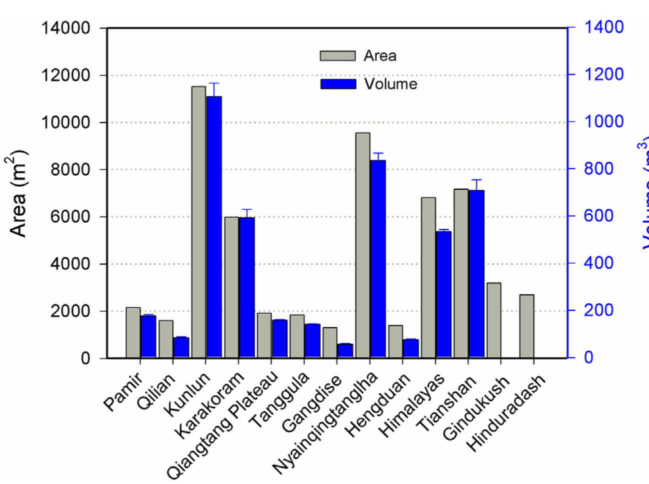

Fig. 3. Distribution of glacial area and volume in the TP and surrounding regions from Yao et al. (2008), Dyurgerov et al. (2002), and Liu et al. (2015).

The Tianshan Mountains of China have suffered a substantial glacier mass loss, caused by higher ablation than accumulation, over the past 50 years. The glacier in the Urumqi River source region persistently shrank from 1962 to 2003. The cumulated mass balance was −10,032 mm, equivalent to 20% of the glacier volume (Ye et al., 2005). During 1962–2009, an ice volume of 29.51×10^6 m^3 has decreased in Urumqi Glacier No. 1, corresponding to a cumulative ice thickness loss of 8.9 m and a mean annual loss of 0.2 m (Wang et al., 2014). The overall loss in total glacier area and mass in the Tianshan Mountains has recently been estimated to be 18 ± 6% and 27 ± 15% from 1961 to 2012, respectively, which are equal to a total area loss of 2960 ± 1030 km^2 and an average glacier mass-change rate of -5.4 ± 2.8 Gt yr^{-1} (Farinotti et al., 2015). A study of glacier changes in Xinjiang by Li et al. (2010b) showed that the total area dominated by the 1800 glaciers retreated by 11.7%, with an average loss of 0.243 km^2. The glacier terminus shrank by 5.8 m a^{-1} during the past 26 to 44 years, a mean retreat rate of 3.5–10.5 m a^{-1}.

3.1.3. Glacier simulations

Numerical simulation and forecast of glacier changes are the frontier of Cryospheric Sciences (Duan et al., 2012; Hock et al., 2017). The distribution of global glaciers is generally discontinuous and scattered. In the field of glacier simulation and forecast, the area of glaciers is smaller than that of a grid point of the GCMs (Global Circulation Models) and is often ignored. Currently, the main progress in simulation and forecast is obtained from studies of single glaciers based on high resolution RCMs coupled with glacier dynamic processes. Therefore, the forecast of the whole change of the global glaciers is still in its infancy. Among the existing regional models, a glacier is often defined using a very simple glacier surface process and is treated as a fixed boundary condition. To solve these problems, a subgrid parameterization scheme has been introduced into a land-surface model coupled with regional models (Jacob, 2001). Lately, an increasing number of

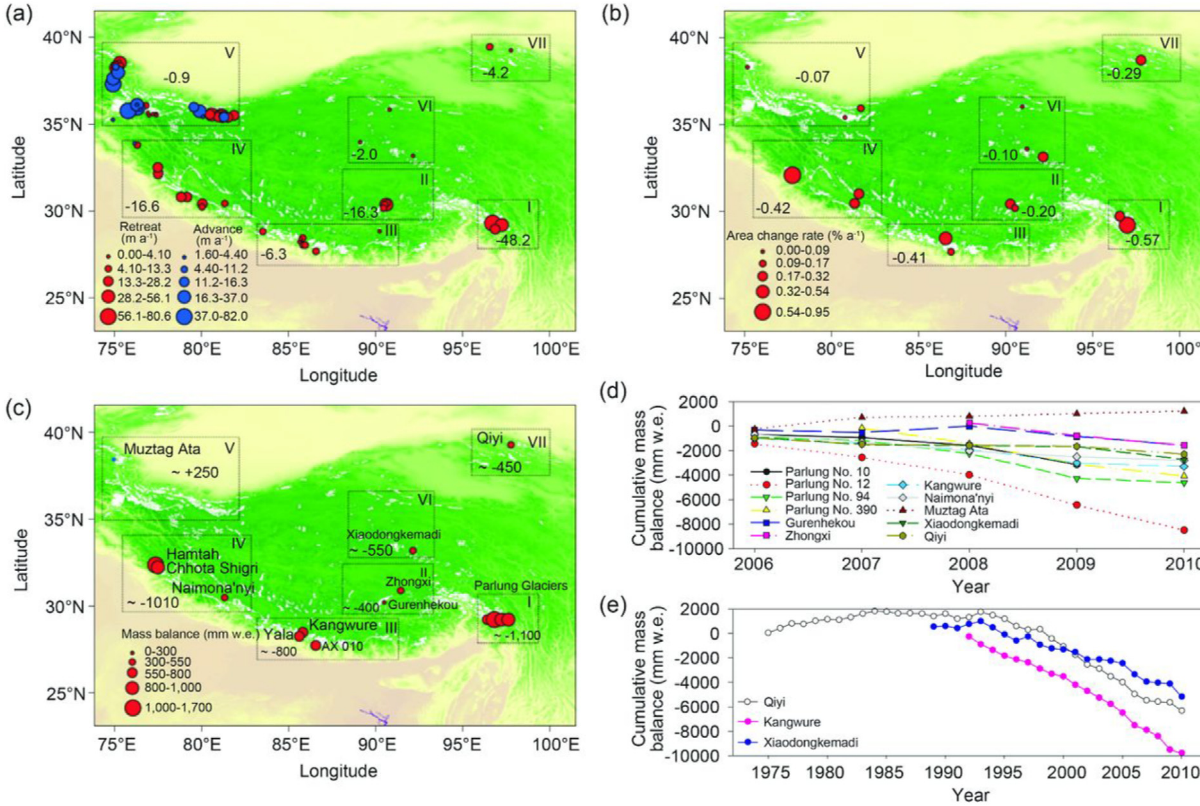

Fig. 4. Spatial and temporal patterns of glacier change in the TP and surroundings. (a) glacier length, (b) glacier area, (c) glacier mass balance, (d) cumulative mass balance for 11 glaciers during 2006–2010, (e) cumulative mass balance for the three longest time-series of glacier mass balance measurements (Yao et al., 2012b).

regional models, namely, dynamic downscaling methods have been applied to investigate climate and atmospheric circulation in alpine glaciers over a large area. For instance, Collier et al. (2013) utilized a coupled high-resolution mesoscale atmospheric (WRF) and physically-based climatic mass balance (CMB) modeling system (including the feedback of glacial mass balance to the atmosphere), WRF-CMB, to analyze the response of glaciers to climate combined with remote sensing data and in-site glaciological and meteorological observations in the Karakorum Mountains during the ablation season of 2004. They found that interactive coupling has a localized but noticeable influence on the near-surface meteorological forcing data, and incorporation of CMB processes heightens the simulation of land-surface temperature and snow albedo. Moreover, the feedback from the glacial model exerts noticeable control on the simulated mass balance, decreasing the ablation simulation. The interactively coupled model displays a new and multi-scaled tool to solve interaction processes between mass balance of the mountain glaciers and atmosphere. Mölg et al. (2012, 2014) explored the influence of the summer monsoon and mid-latitude westerlies on glaciers by groundbreaking use of WRF and glacier mass balance.

A glacier dynamical model has introduced all kinds of dynamical parameters and can commendably reflect the influence of climate change on glaciers (such as glacial temperature and terminus movement), and the forecast results are thus comparatively accurate (Li et al., 2007). The limited studies on glacial dynamics in China are about the frequency response of Glacier No.1 (Wang and Liu, 1984), the simple simulation of the main line thickness (Cao and Meier, 1987), and some theoretical discussion regarding the glacier in the Yili River valley, Xinjiang Autonomous Region (Ye et al., 2003). Beyond these, other related research is relatively scarce. Wang and Liu (1984) predicted glacial changes according to the theory of frequency response and showed that the magnitude of glacier recession might be the greatest in the last 30 years approximately 2006, when the glacier length shortened by 14%. Cao and Meier (1987) calculated longitudinal profile parameters when glaciers reached a stable state using a statistical correlation method and predicted the area and volume of Glacier No.1 to be 86.3% and 60.5% in 1980 respectively. Glacier No.1 would be divided into two branches in 1993. The correctness of this prediction was partly verified later. Ye et al. (2003) investigated the response of alpine glacier area and runoff in the Yili River to climate change using the glacier ice-flow model. They indicated that the sensitivity of glaciers to climate depends on glacier area, and the change of glacial runoff was out of sync with climate, depending on the glacier area and the rate of atmospheric warming. Xie et al. (2005) predicted the changes of runoff, area, volume, and ELA for the glaciers in every large basin in western China in light of the data from the Chinese glacier inventory and the glacier system model under several climatic scenarios of varying warming rates. On average, if climate warming proceeds, the glacier runoff will initially increase, reach a climax by the year of 2030 and decrease thereafter. The glacier area in China will shrink continuously by about 6%–9% by 2030 and by 10%–15% by 2050. Considering their strong physical process analytical ability, the dynamical models play an important role in predicting glacier changes. Among those models, the "frequency response model", "section-factor model", and "ice flow model" are relatively robust and capable of simulating and analyzing alpine glaciers (Li et al., 2007). A 3-D thermodynamic model simulation (GLIMMER, Genie Land Ice Model with Multiply Enable Regions) performed by Wang et al. (2009) matches well with the distribution of observed present day glaciers (1951–2000) in the TP, yet the simulated glacier areas at the Last Glacial Maximum (LGM, ~18,000 years ago) and Little Ice Age (about 400–500 years ago) are larger than the observed data, and the projected glacier area will decrease in the 21st century, but the glaciers will advance twice in 2045 and 2080 under SRES scenarios.

Duan et al. (2012) researched the Urumqi Glacier No.1 in combination with the glacial flow simulation and mass balance model by

solving a two-dimensional nonlinear Stokes equation. According to their predictions, under a scenario of 0.17 °C decade^{-1} warming rate and unchanged precipitation in the region from 2010 to 2070, when considering the physical process of glacier change, the terminus will retreat slowly before 2040, but the thickness in the ablation area will vary greatly, after that, the terminus will retreat rapidly in 2070. The theory of glacier flow has proven that the retreat of alpine glaciers will be dramatic under future global warming. Glacier temperature and flow velocity are essential parameters for the glacier dynamical model (Li et al., 2007) and are also decisive variables for discerning glacier response to climate change. The two-dimensional higher-order thermomechanical flowband model reproduced well the surface-flow velocity and glacier temperature of the east Rongbuk Glacier, Mount Everest (Zhang et al., 2013). Zhao et al. (2014) employed a three-dimensional, thermomechanically coupled full-Stokes model to simulate the evolution of the Gurenhekou glacier, southern TP. Their results show that the glacier retreated at an average rate of 8.3 m a^{-1} when summer air temperature increased at 0.02 K a^{-1} over the past 50 years. A predominant warming rate of 0.05 K a^{-1} over the TP during the 21st century will cause about a 0.9 ± 0.3% a^{-1} loss of glacier area and 1.1 ± 0.6% a^{-1} shrinkage of glacier volume for the period 2008–2057.

3.2. Research on snow cover

Snow cover is an extremely sensitive element of the cryosphere. Snow cover is solid precipitation and is double-influenced by precipitation and temperature. Meanwhile, snow cover, with high albedo (0.8–0.9 for fresh snow), low thermal conductivity and roughness length, strongly and directly provides feedback to atmospheric circulation. The thickness and melting of snow cover can affect soil temperature, soil freezing and thawing processes (Zhang, 2005; Yang et al., 2008) and soil moisture, which in turn affect water and biochemistry cycles (Seneviratne et al., 2010). Snow cover is viewed as a regulator to spring drought in arid regions since its change in the TP strongly influences spring runoff (Qin et al., 2006b). It has been confirmed that snow cover affects climatic fluctuation in China (Qian et al., 2003; Wu and Qian, 2010) and spring phenology (Wang et al., 2013a). In short, the part of that snow plays in the climate system contains strong positive feedbacks linked to albedo, and indirect responses linked to insulation of the frozen ground, moisture storage, and latent heat.

3.2.1. Monitoring of snow cover

Several data sources for the TP snow cover are available: ground meteorological stations, satellite observation by NOAA, SMMR (Scanning Multichannel Microwave Radiometer) by NASA, and EOS/MODIS (Earth Observation System/Moderate Resolution Imaging Spectroradiometer) satellite retrieval products (Qin et al., 2006b; Wang et al., 2007; Che et al., 2008; Zhang et al., 2008, 2012; Tang et al., 2013; Dai et al., 2017). There are > 200 meteorological stations in western China, but not in the western TP. The network in the TP is composed of 60 primary stations, of which 35 are located in Qinghai province and 25 in the Tibetan Autonomous Region (Xiao et al., 2007). The site data is discontinuous in time-series and the distribution of stations is inhomogeneous. In the mountains with extensive snow, especially in the western TP, there are few or even no stations. Thus, the regional representativeness of site data is quite limited. The accuracy and integrated information of snow cover in the TP cannot be gathered from only the station data. Satellite remote sensing data can serve the purpose, but most of the satellite data are often disturbed by clouds and likely to be missing in high mountains (Zhu and Woodcock, 2014). They can be used simply to identify snow cover on the hemispheric or continental scale due to low horizontal resolution. The station data in the TP cannot perfectly describe real surface snow cover. Although the time series of stations and satellites are similar, such as heavy snow cover in 1977/78 and 1988/89 and light snow cover in 1984/85, differences still exist between them. An apparent example was a very heavy snow in 1985/86, which was recorded by SMMR and NOAA but not by stations (Qin et al., 2006b). Compared with NOAA, EOS/MODIS, with higher resolution in the satellite spectrum, as well as space and time, is also more powerful in distinguishing clouds and snow. It can monitor the snow cover well in forest areas. As a result, MODIS data has a greater potential for monitoring snow cover change in regions with complex terrain (Wang et al., 2007; Dai et al., 2017). In the meantime, analysis of MODIS data can be used to discuss the distribution and seasonal changes of snow cover in the TP. However, MODIS data has a short time span, being available since February 2000, so to examine the long term trend of snow cover changes, station data is still the first choice. Therefore, for a concrete issue, such as spatial distribution, seasonal changes, and annual fluctuation of snow cover, combining several kinds of data should be the optimal option.

3.2.2. Snow cover change

There are three high value centers in snow cover from the perspective of stations: the southern high value center located on the north slope of the Himalayas; the eastern part of the Tanggula and NyenchenTanglha Mountains; and the Amne Machin and Bayan Har Mountains in the eastern TP (Wei et al., 2002; Xu et al., 2017). A high-resolution satellite snow cover product shows that high snow coverage is distributed in high and large mountains. Most of the stable snow cover is in the southern and western TP, where the rainfall is generated by the Indian monsoon climbing over the TP (Fig. 5). By contrast, relatively less snow remains in the central TP despite of an average altitude of over 4000 m because of the topography blocking effect of the Himalayas and Karakorum Mountains (Pu et al., 2007).

The extent of snow cover monotonously increases with altitude. Over 5500 m, the annual extent of snow cover is > 30%, the maximum value appears in spring and autumn, the secondary maximum appears in winter and the minimum appears in summer (Fig. 6a). Obvious seasonal changes in snow cover extent emerge on the whole TP (Tang et al., 2013, Fig. 6b). Snow cover over the TP occurs from October to May, marginally in September and June, and is less or basically nonexistent in July and August. Snow cover in the eastern plateau has the most remarkable annual change, dominating the annual change on the whole plateau, contrary to the phase of annual fluctuation in the western heavy snow regions (Tang et al., 2013). Autumn and winter snow cover in the northern TP increase rapidly but melt slowly in the next spring (March–May) (Zhang et al., 2008).

Snow cover in the TP shows a decreasing trend, albeit with a large interdecadal variability. Strictly speaking, it increased significantly from the 1960s to the late 1980s, but decreased in the 1990s (Wei et al., 2003; Zhang et al., 2008) and the 2000s (Deng et al., 2017). Shen et al. (2011) pointed out that snow cover in the TP shows evident interdecadal variations, undergoing drastic transformations in the 1840s, the 1880s, the 1920s, and the 1960s based on ice cores and observed snow cover. Snow cover in the whole TP has decreased at a rate of 4.0% decade^{-1} from 1997 to 2011 (Shen et al., 2014, Fig. 7d). The trend analysis of snow cover days showed that 34.14% of the study area had a decreasing tendency during 2001–2011, with significant 5.56% decreases. Distinct decreasing chiefly appears in high elevation areas (esp. in the Hengduan Mountains and northern Karakorum-Kunlun mountains). However, 24.75% of the study area shows an increasing trend, with significant increases of only 3.9%. Most of these regions are in the southern Karakorum-Kunlun mountains, the northern Qilian mountains and the central TP (Fig. 7a–b) (Tang et al., 2013). There is also a decreasing tendency for the snow coverage rate (from January to April) in most parts of the TP (Wang et al., 2013a, Fig. 7c). The extent of snow cover decreases in winter and spring and increases in summer and autumn connected with temperature and precipitation. In winter, changes of snow cover extent are more susceptible to precipitation whereas in summer temperature becomes a crucial factor (Wang et al., 2007). The regional mean depth and days of snow cover increased at the rate of 0.32 mm decade^{-1} and 0.40 d decade^{-1} from 1961 to 1990

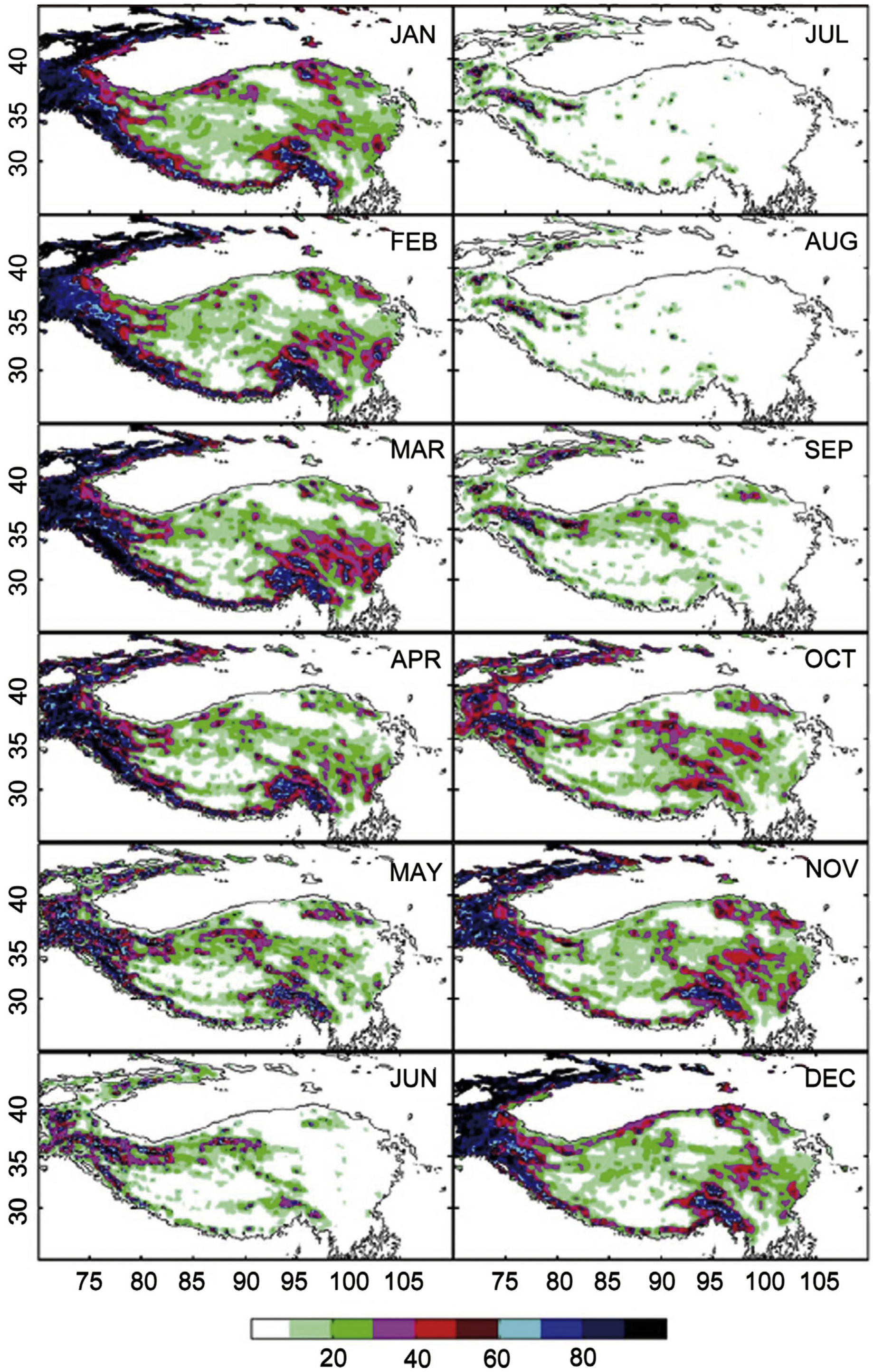

Fig. 5. Distribution of monthly mean snow cover fraction (%) in the TP across the whole year (Pu et al., 2007).

but decreased at the rate of 1.80 mm decade^{-1} and 1.59 d decade^{-1} from 1991 to 2005 (You et al., 2011). Similar variation trends with regard to the snow depth and the number of snow-cover days in the TP were acquired by Xu et al. (2017) during 1961–2010, meanwhile the overall duration of snow cover has significantly decreased at the rate of 23.5 ± 1.2 d decade^{-1}, which is collectively determined by delayed starting and advanced ending times.

3.2.3. Snow cover simulations

Restricted by the shortage of observational data, studies on the interaction between changes of snow cover on the TP and climate in China are still few. Most of them start from GCMs (Zhang and Tao, 2001; Rangwala et al., 2010; Wei and Dong, 2015) and regional models are seldom used in early research. For example, Rangwala et al. (2010) employed the GISS-AOM (Goddard Institute for Space Studies-Atmosphere Ocean Model, NASA) GCM to elucidate the mechanism of snow cover for a large warming trend over the TP. Based on a series of equilibrium GCM simulations, Qian et al. (2011) provided a striking insight on the importance of light-absorbing aerosols (LAA, e.g., black carbon (BC) and dust) in the TP snowpack. LAA can decrease surface reflection such as surface darkening, which further decreases snow albedo and accelerates snow melt. Such feedback induces a greater

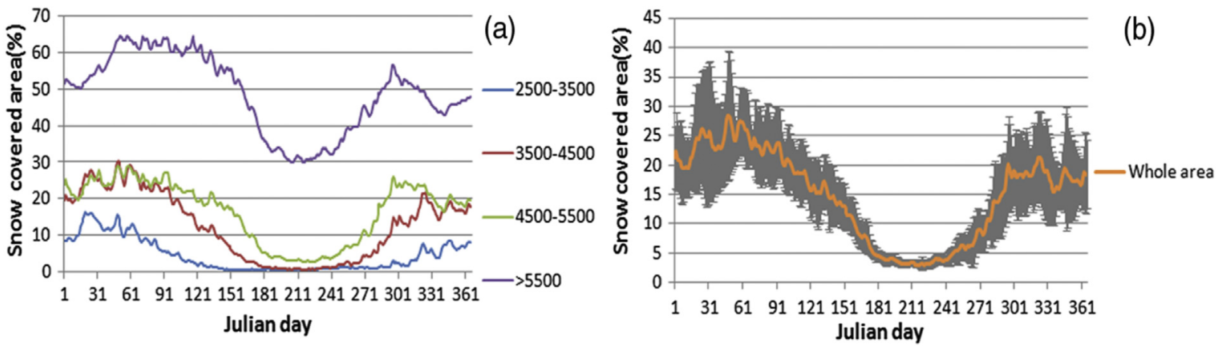

Fig. 6. (a) Annual cycle of snow cover (%) for four different elevation zones and (b) the whole area of the TP during 2001 to 2011. The error bars in Fig. 6b show the standard deviation of snow cover from 2001 to 2011 (Tang et al., 2013).

surface air temperature and snow melt efficacy, and thereafter affects the Asian monsoon climate and hydrologic cycle (Qian et al., 2011, 2015). The modeling results of an ocean–atmosphere GCM by Xu et al. (2016) confirmed BC deposited onto the surface of snow or ice together with CO2 caused snow retreating trends on the TP. BC increase in snow largely contributes to the intense elevation dependency of surface warming over the TP. Yet the GCMs with low horizontal resolution cannot describe the meso-and microscale topography and land-surface characteristics well and the forces of other factors on regional climate change. By contrast, high resolution regional models containing a chain of suitable parameterization schemes and physical processes can better reveal the physical mechanisms of regional weather and climate (Gao et al., 2011). Simulations by regional climate model NCAR RegCM2 showed that the abnormal snow depth and cover in the TP clearly influence atmospheric circulation (Qian et al., 2003; Liu et al., 2005). RegCM series have great ability to simulate days of snow cover, snow depth, and the beginning and ending time of snow cover (Shi et al., 2011; Ji and Kang, 2013). In the 21st century the days and depth of snow cover will both decrease, the starting date will be delayed and the ending date will be advanced (Shi et al., 2011). The RCM (Model Atmospheric Regional, MAR) reasonably simulates the extent and duration of snow cover in the Himalayas under real climate conditions (Ménégoz et al., 2013). There are considerable differences in the simulation results from within and outside China, and some even have fundamental divergences (Zhang and Tao, 2001). The main reasons are that the surface hydrologic process and treatment of snow forcing are too simple, suffering from the restriction of snow data, and the design schemes of snow forcing are different (Zhang and Tao, 2001).

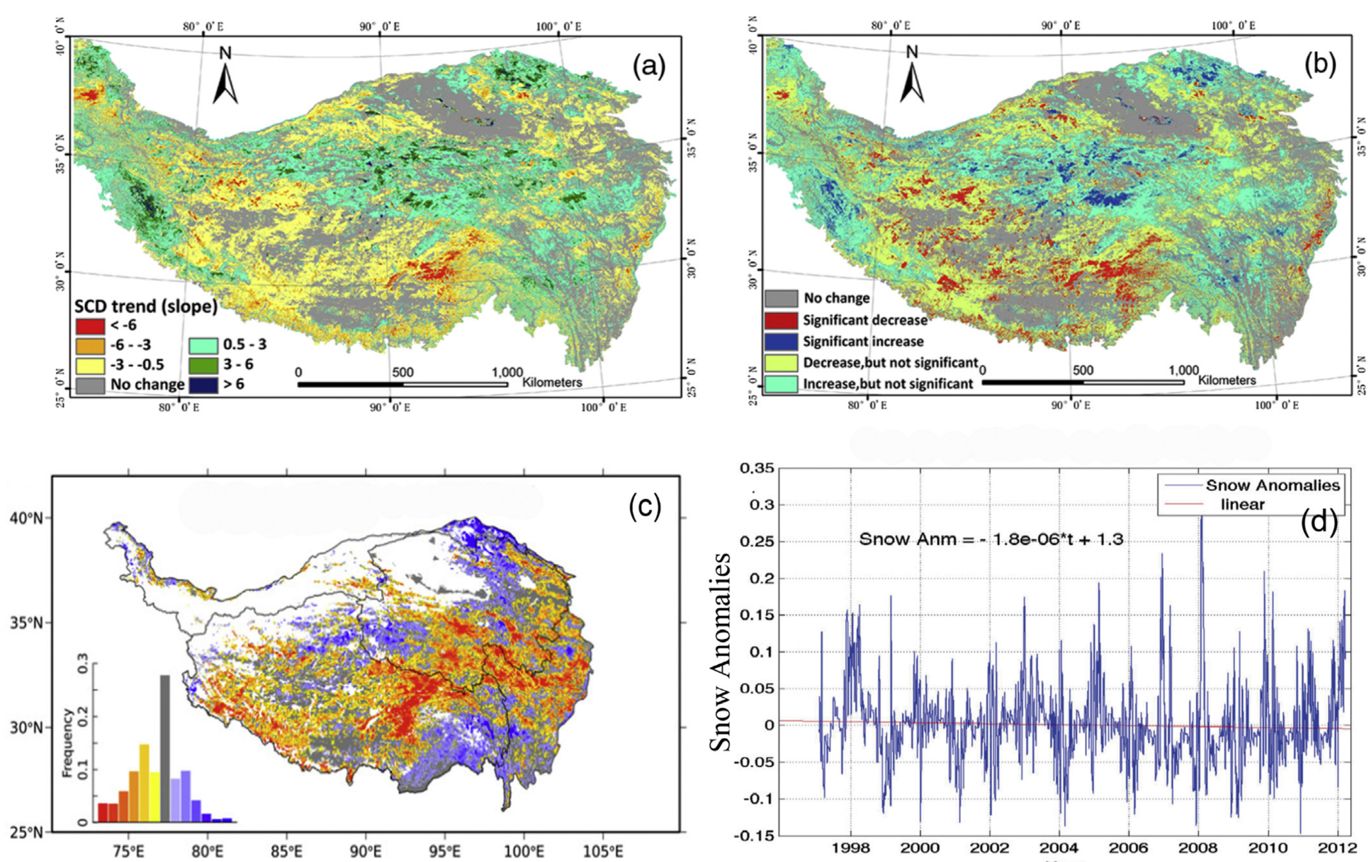

Fig. 7. (a) Spatial trend in snow cover days (SCD) from 2001 to 2011 and (b) significance of the trend. Trends are termed significant for pixels in which $p < .05$ (Tang et al., 2013); (c) Trend in snow-cover fraction (%) during January–April between 2000 and 2011 (Wang T et al., 2013); (d) Daily anomalies of snow cover between February 4, 1997 and March 15, 2012 (Shen et al., 2014).

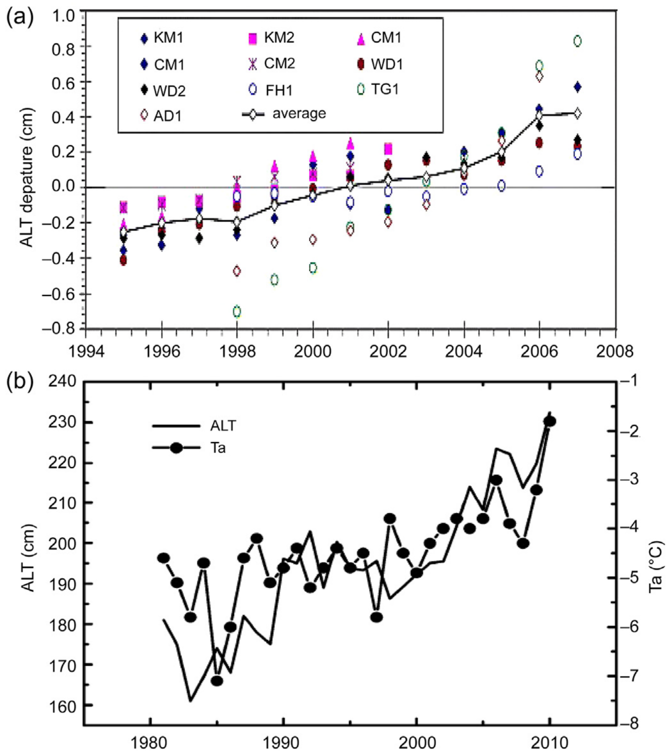

Fig. 8. Variations of the ALT departure from its mean for each site along the Qinghai-Tibet Highway (QTH) over its period of record (Wu and Zhang, 2010) and (b) ALT and air temperature (Ta) along the QTH from 1998 to 2010 (Li et al., 2012).

No consensus has yet been reached on whether the main climate effects of snow cover are radiation or hydrology. Barnett et al. (1988) argued that the single reflection effect for snow is not obvious, however, combined with snowmelt and evaporation it clearly attenuated the intensity of monsoon circulation. Vernekar et al. (1995) indicated that the weak summer monsoon results from energy being consumed for the ablation of more snow cover, which leads to lower ground temperature and sensible heat flux, decreasing the meridional temperature gradient. Simulation results of RegCM2 by Liu et al. (2004) realized the reflection effect is primary, while the effects of snowmelt and evaporation are relatively small. Additionally, snow cover models substantially overestimate the depth and days of snow cover in the TP because they neglect the impact of wind speed. When underscoring this, the snow cover model can estimate reasonable snow cover dynamics and depth to a great degree (Yuan et al., 2016).

3.3. Research on the change of frozen ground

Frozen ground is an indicator of climate change. Large areas of permafrost and SFG are one of the most obvious features on the surface of the TP. Freezing-thawing processes are affected by annual climate change. Meanwhile, changes in soil temperature and moisture can significantly affect the local and East Asian earth-atmosphere system through exchanges of water and energy between the land surface and atmosphere (Yang et al., 2007c). Freeze-thaw cycles can strengthen the heat exchanges between the land surface and atmosphere and the TP monsoon (Cheng and Wu, 2007). Soil carbon already stored in frozen ground could be released into the atmosphere by the cycles (Zhang, 2007) and permafrost degradation (Koven et al., 2011; Schuur et al., 2015; Ding et al., 2017). The observation sites for monitoring frozen ground in the TP are mostly concentrated along the Qinghai-Tibet Highway (QTH) (Wu and Zhang, 2010; Wu et al., 2012b; Yang et al., 2010; Zhao et al., 2010), and a dozen permafrost drills at high altitude are available along the Qingkang Highway in the Bayan Har Mountains (Luo et al., 2019). Many scholars have done a large amount of research on the changes of frozen ground using different data.

3.3.1. Changes of frozen ground

Permafrost temperatures in the TP decline as elevation increases (Zhao et al., 2010) and have generally exhibited an increasing trend against the background of climate warming over the last decades. The increase of mean annual ground temperature (MAGT) in permafrost regions is more distinct than mean annual air temperature on the TP (Wu et al., 2013). This is caused, in part, by the decreased snow cover in recent years as stated above, which allows surface absorption of more incoming solar radiation to warm the permafrost. Another potential reason for the increased difference of ground-air temperature could be the increase in downward longwave radiation influenced by an increase in surface specific humidity (Rangwala et al., 2010). The increasing rate of cold permafrost (MAGT < -1.0 °C) is higher than warm permafrost (MAGT > -1.0 °C) (Wu et al., 2012b). The permafrost temperature at 6 m depth in the QTH increases obviously. From 1996 to 2006, the mean annual temperature at 6 m depth increased by 0.12–0.67 °C (0.43 °C on average) (Wu and Zhang, 2008), however, the increasing trend has begun to slow in recent years, at a rate of 0.02 °C a^{-1} from 2006 to 2010 (Wu et al., 2012b).

With temperature continuously rising on the TP, permafrost has experienced a regional degradation trend. According to the island permafrost maps (1:100000) compiled in 1975, there were 65 km^2 of permafrost out of the total 320 km^2 in the southern QTH. Comprehensive field investigation in 1996 showed the permafrost area has decreased to 42 km^2, the most pronounced permafrost degradation occurred in the mountain regions and plains (Jin et al., 2000). Permafrost degradation has occurred on a large scale in the northern lower limit of permafrost regions (Xidatan) over the last 30 years, decreasing from 160.5 km^2 in 1975 to 141.0 km^2 in 2002 (Nan et al., 2003; Wu et al., 2005). The area of permafrost in the TP has reduced from about 1.50×10^6 km^2 in 1975 to about 1.26×10^6 km^2 in 2006 (Jin et al., 2011) to the more recent value of 1.06×10^6 km^2 over the period 2003–2012 (Zou et al., 2017). The northern lower limit moved southwards by 3 km during 1975–1996 (Jin et al., 2000). Detection radar in Xidatan, near the northern lower limit of permafrost distribution on the TP, shows that the lowest permafrost elevation was 4385 m, and it has risen by 25 m since 1975 (Wu et al., 2005). The southern lower limit of permafrost along the QTH moved northwards by 10 km during 1975–1996 (Jin et al., 2000), and the lower altitudinal limit in the south rose by 50–80 m in the past 25 years (1996–2001) (Cheng and Wu, 2007; Yang et al., 2010).

The upper permafrost is the active layer, which is subject to seasonal freezing and thawing. The active layer thickness (ALT) in the QTH from the 1980s to the early 1990s increased by 1 m (Wu and Tong, 1995).The average thickness increase in high mountains was about 0.24–0.5 m during 1995–2000, in the plain areas it was about 0.05–0.39 m, and in the mid-and lower-mountains it was about 0.18–0.4 m (Wu and Liu, 2004). The mean ALT in the QTH increased at a rate of 7.5 cm a^{-1} during 1995–2007 (Wu and Zhang, 2010; Fig. 8a). Because of QTH construction, most of the sites buried at a relatively shallow depth (about 6–8 m) were stopped. In 2005, the drill depth of 27 permafrost monitoring sites built along the new QTH reached over 15 m. From 2006 through 2010, the permafrost ALT increased by 6.3 cm a^{-1} (Wu et al., 2012b). A study of Li et al. (2012) suggested that the permafrost ALT along the QTH increased at a rate of 1.33 cm a^{-1} during 1998–2010 (Fig. 8b), which is lower than that of Wu and Zhang (2010) mainly because the ALT is less affected by road engineering and human activities. Additionally, the difference between the study time-periods may be another reason. Overall, the ALT becomes thicker, and namely, permafrost tends to thin with ground temperature increase in the TP.

The highest values of freeze-thaw cycles across China are in the TP,

and are largely controlled by altitude (Peng et al., 2016). The seasonal soil freeze depth (SFD) at most sites on the TP are > 2.4 m and increase with latitude and elevation (Peng et al., 2017). Entering the 1990s, the thickness of the SFG layer began to thin in the northeastern, southeastern and southern TP. The maximum SFDs at representative stations in these areas decreased by 0.02 m, 0.05 m, and 0.14 m compared with the 1980s (Wang et al., 2001). In the Yellow River source region, the maximum SFD decreased by an average of 0.12 m between 1981 and 2008 (Jin et al., 2010). Differently, the maximum SFDs have thickened in the Qaidam Basin and central TP since the 1990s (Wang et al., 2001). Zhao et al. (2004) indicated that distinct changes of the SFG appeared in regions with the thickest SFG such as the central TP, whereas the smallest change occurred in the southeastern TP. The duration of SFD has shortened by over 20 days in the northeastern and interior TP for the 1967–1997 period but has remained relatively stable in the northwestern and southeastern TP. Regional average SFD has decreased significantly, at 0.22 ± 0.004 cm yr^{-1} from 1950 to 2009 (Peng et al., 2017).

3.3.2. Frozen ground simulations

Numerical simulation is an indispensable tool for investigating and forecasting frozen ground changes. The results of a one dimension heat transfer model with phase change showed the average active layers deepened by over 0.3 m in most regions of the northern TP during 1980 to 2001, and in some regions even have deepened by 0.9 m (Oelke and Zhang, 2007). Simulations of a response model that uses data from a digital elevation model and air temperature showed that the permafrost area continuously decreased from 1.60×10^6 km^2 in the 1960s to 1.27×10^6 km^2 in the 2000s, the permafrost loss started to accelerate in the 1980s, with the degradation area in the 2000s reaching 1/5 of the total area in the 1960s (Fig. 9). The constant permafrost regions were primarily distributed in the Qilian Mountains and the central and western TP, while the degradation regions were distributed in the margins of the permafrost and SFG regions (Cheng et al., 2012). Li et al. (1996) used numerical methods to simulate frozen ground changes in the TP under a climate warming rate of 0.04 °C a^{-1} and suggested that < 10 m thick permafrost regions will degrade into SFG regions after 50 years. According to the elevation and freezing index models, permafrost in the TP will essentially not change in the future 20~50a. However, after 2009 when the annual temperature rises by 2.91 °C on average, permafrost will be reduced by 58.18% (Li and Cheng, 1999). Nan et al. (2005) predicted the TP permafrost changes in the future 50a and 100a by revising the models of Li et al. (1996). The permafrost area will decrease by about 8.8% and the regions with high temperature of $T_{cp} > -0.11$ °C will degrade in 50a when annual temperature warming is 0.02 °C. In 100a, under the same conditions, the permafrost area will decrease by about 13.4% and the regions where the $T_{cp} > -0.5$ °C will degrade. If the rate of annual warming is 0.052 °C a^{-1}, the permafrost in the TP will degrade by 13.5% in the future 50a and 46% in the future 100a. The regions where the $T_{cp} > -2$ °C will degrade into SFG or unfrozen ground regions. Guo and Wang (2013) adopted the Community Land Model (CLM4) driven by RegCM3 to simulate the area and distribution of permafrost and SFG in the TP. The results show that the area of near-surface permafrost was 151.50×10^4 km^2 for 1981–2010, which was larger than the previous observation (111.80–150.0 × 10^4 km^2). Under the warming rate of 0.44 °C $decade^{-1}$ on the TP (1981–2010), the area of permafrost decreased at a rate of 9.20×10^4 km^2 $decade^{-1}$, the ALT increased at a rate of 0.15 m $decade^{-1}$, and the maximum SFD of SFG decreased at a rate of 0.34 m $decade^{-1}$. At 1 m depth, the duration of freezing in permafrost and SFG shortened by 9.7 and 8.6 d $decade^{-1}$, respectively, due to later onset and earlier ending of soil freezing. In the case of future global warming, predictions from the CLM4 showed that the area of near surface permafrost will continue to decrease in the 21st century, at a rate of 9.9×10^4 km^2 a^{-1} (Fig. 10), and the deepening trend in the ALT will keep going, reaching 1.5–2.0 m in 2030–2050 and up to 2.0–3.5 m in 2080–2100 (Guo et al., 2012). Relying on a GCM emission scenario, a simulation based on Kudryavtsev's equations showed that the ALT will increase by 0.1–0.7 m in 2049 and 0.3–1.2 m in 2099, the greatest increase being in the northwestern and southwestern TP on account of high ground temperature and low moisture (Pang et al., 2012).

4. Interactions between cryosphere in the TP and climate

The cryosphere elements in the TP have noticeably changed under global warming. To address the issue of influencing modes and processes of these changes on climate over the TP and its vicinity, conducting research on cryosphere-atmosphere interactions is a hugely required and effective method (Yang et al., 2012). Research on the interaction between the TP cryosphere and climate started early but was limited by the complexity of the glacier change process, snow cover process, and frozen ground hydrothermal processes, and invariably focused on how climate change influences the cryosphere. How the cryosphere elements influenced the climate on the TP, especially their feedbacks to climate in surrounding areas, have been scarcely explored till now. The uneven and low-density distribution of meteorological stations impedes our ability to deeply comprehend the TP climate change. Meteorological stations over high altitude regions are very sparse, and most are situated in valleys and low altitudes (Fig. 1b). Therefore, the site data cannot fully represent the TP climate. Beyond that, site data with short time-period (< 100 years) cannot be used to assess the long-term climate trend (Yang et al., 2010). The restrictions from meteorological stations and site data make it difficult to achieve the mechanism of cryosphere change in high-altitude mountain regions (Pepin et al., 2015). Remote sensing data, albeit with relatively high resolution, has the limitation of short time span and excludes upper atmosphere data corresponding to the cryosphere, which can hardly identify atmospheric and climate processes that cause the changes of observed cryosphere elements (Wang et al., 2018a).

The global reanalysis data series with low resolution imposes restrictions on the representativeness in mountain regions. All climatic factors are prone to be affected by topography, and precipitation is the most complex one. Utilizing the reanalysis grid data to represent the TP precipitation is perhaps not appropriate (Wang et al., 2018a). The land-atmosphere coupled model (i.e., dynamic downscaling method) is a valuable tool to investigate the interactions between land and atmosphere (Buteau et al., 2004). It has provided an approach to quantifying the influences of cryosphere changes on regional climate. The regional climate model (RCM), a kind of dynamic downscaling method, can provide more detailed regional information and has already been applied on the TP. Simulations of climate based on the RegCM series over the TP represent the basic features of temperature and precipitation and partly fill the pronounced high-altitude observational gap (e.g., Gao et al., 2011; Wang et al., 2013d, 2014b, 2016). WRF can successfully simulate the frequency of precipitation and orographic rainfall (Maussion et al., 2014), and annual cycles and temporal trends of temperature and precipitation in the wet season on the TP (Gao et al., 2015b). These climate models generally produce some errors such as the overestimation of precipitation and underestimation of temperature when compared with observations, stemming primarily from the improper cumulus parameterization and microphysical precipitation schemes (Yu et al., 2015), together with land-surface models that do not satisfactorily consider specific topography and land-surface processes (Wang et al., 2013d, 2014b).

Soil is the leading part of land-surface processes, and soil freezing and melting states are important parameters. Frozen ground, the product of climate, not only influences regional climate by means of partitioning of surface energy flux but also alters surface runoff and soil infiltration by changing transmissibility and the soil water capacity. Such changes can directly impact hydrologic processes, modify the land surface status, and further influence climate (Yang et al., 2005, 2007b;

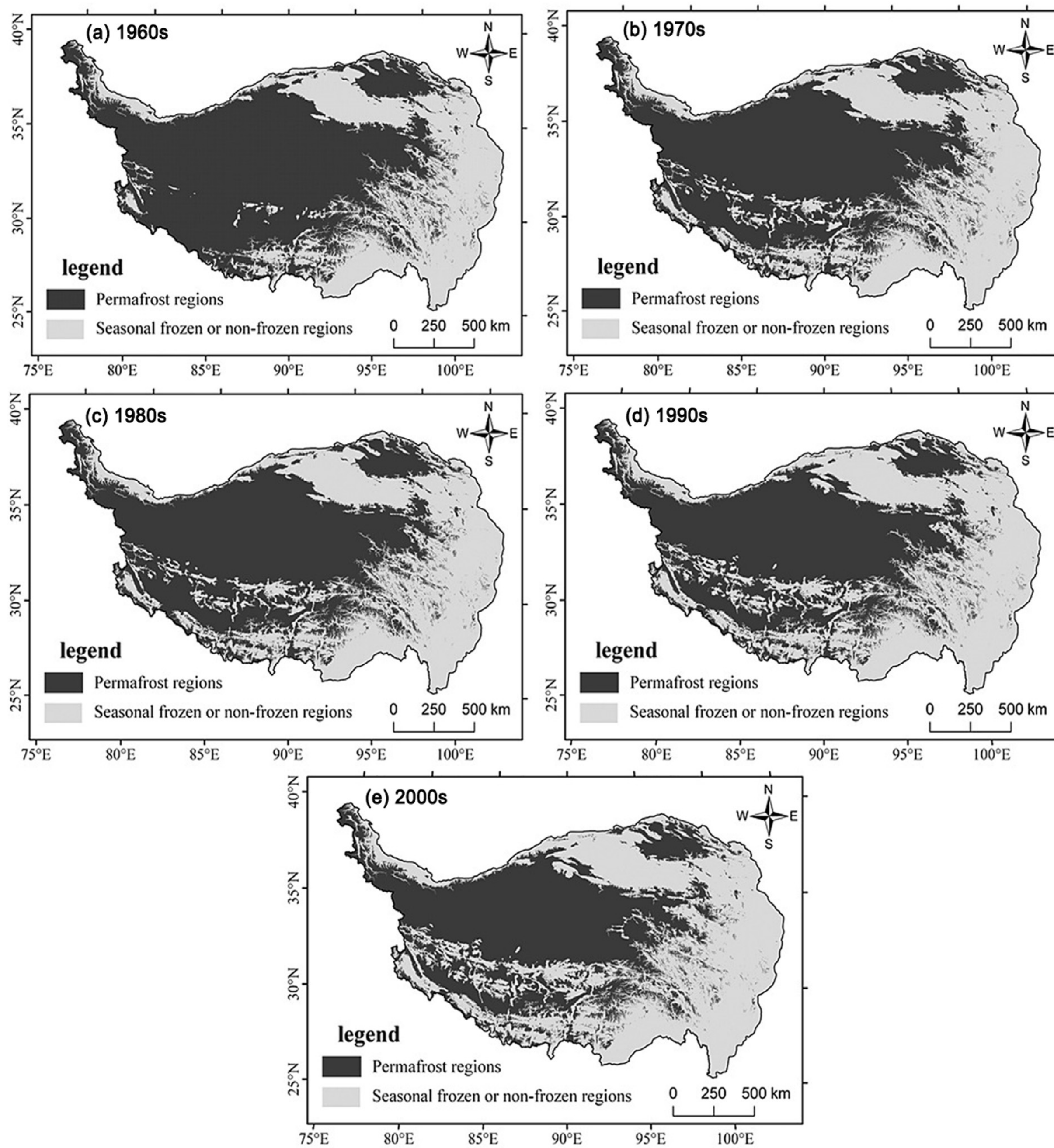

Fig. 9. Simulated permafrost distribution in the 1960s, 1970s, 1980s, 1990s, and 2000s in the TP during 1960–2009 (Cheng et al., 2012).

Wang et al., 2016). Climate change can also be the significant driver for soil freezing and thawing processes (Henry, 2008; Peng et al., 2016). Yang et al. (2003, 2007c, 2007d) comprehensively investigated the soil freezing-thawing processes and their impacts on the energy and water cycles based on the China-Japan international cooperation projects (GAME-Tibet, 1996–2000 and CEOP/CAMP-Tibet, 2001–2005). With a large spatial variability, the freezing-thawing processes alter soil heat-absorbing and releasing processes and enhance heat exchange in the land-atmosphere system (Zhao et al., 2004; Yang et al., 2007d). The sensible and latent heat flux clearly responds to soil temperature and moisture under different freezing and melting states (Guo et al., 2011). Some upgraded parameterization schemes have already been developed and applied in the TP (Li and Koike, 2003; Yang et al., 2005; Zhang et al., 2007; Li and Sun, 2008; Wang et al., 2015a, 2017a), since correct description of the water-heat status in frozen ground is one of the most important causes determining the performance of models. However, owing to the lack of observations and the complexity of frozen ground, research on surface characteristic parameters is still a thorny problem in frozen ground regions. Many scholars qualitatively examined the effects of inhomogeneity or certain land-surface characteristics on land-atmosphere interactions and climate (Yang et al., 2003, 2007b; Chapin et al., 2005; Ma et al., 2008). Quantitative research on the effects of a single surface characteristic (such as soil and snow cover) on the water-heat exchanges between land surface and atmosphere will further contribute to understanding the energy and water cycles. One of the effective methods used in quantitative studies is conducting sensitivity experiments for different surface features expressed by the parameterization schemes in the climate models in order to comparatively analyze the results (Wang et al., 2014b, 2015b). Additionally, many existing works about frozen ground are mostly carried out along the

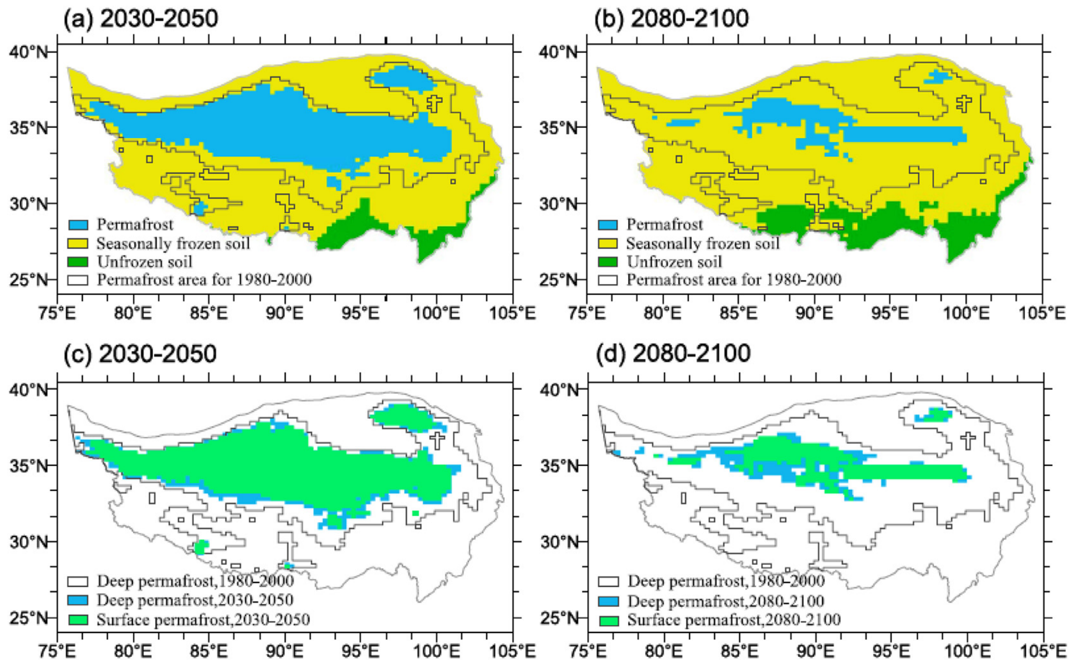

Fig. 10. CLM4 simulated mean near-surface permafrost areas for the period of (a) 2030–2050 and (b) 2080–2100. (c) and (d) same as (a) and (b), but for deep permafrost (soil at depths between 10 and 30 m) (Guo et al., 2012).

QTH. The distribution and dynamic change of frozen ground in most regions of the TP are still unknown, as well as how it responds to climate forcing. Increasing the density and representativeness of the observation network and initiating long-term remote sensing observations such as InSAR are the priority in the TP for the near future (Yang et al., 2010; Daout et al., 2017).

The traditional methods of simulating glacial mass balance are one-way or offline. These methods often hardly consider the feedback from changes of glacial surface conditions to atmospheric forcing. In addition, a glacier is only regarded as a kind of static land type and poorly depicted in atmospheric models (Collier et al., 2013; Kumar et al., 2015). Alternatively, RCMs embedded with an interactive glacier parameterization containing glacier energy and mass balance schemes have been developed to offer a complete simulation of glacier-climate interactions (Kotlarski et al., 2010; Collier et al., 2013; Kumar et al., 2015). Soil freezing and snow are land-surface processes, and their changes should not be isolated (Warrach et al., 2001). Because of the insulation effect, soil temperature in regions with thick snow cover is higher than those without snow cover (Zhang, 2005). Thick snow cover can greatly impact the daily change of soil temperature, thus the change in the daily temperature range can record the condition of surface snow cover. Abnormally heavy snow can influence ALT through controlling ground temperature. Interannual variations in snow depth have great significance to maintaining frozen ground in the TP (Yang et al., 2008). Therefore, separate research on the climate effects of snow cover and frozen ground are far from comprehensive. The inclusion of both factors above will probably help boost the accuracy of climate models (Wang et al., 2017a).

So far, the research framework of Cryospheric Science has been designed, and many issues involved have not been completely addressed. Fortunately, several insightful methodologies and research directions are proposed. For example, the linkage between the cryosphere and other earth spheres is one of the keys and should be emphasized in the future (Qian et al., 2018). Changes in cryosphere elements pose vast effects on the climate system. The temporal and spatial variations of snow cover and glaciers can greatly affect global energy and hydrologic processes through high albedo, as well as modulate climate dynamics at regional and even global scales. A great deal of evidence from observations and simulations has affirmed that the LAA deposition in snow/glaciers has remarkable impact on climate warming and snow/glacier reduction via snow-darkening and radiative forcing (e.g., Qian et al., 2014, 2015; Zhang et al., 2018). Constraints on the sources and amount of LAA deposited on snow/glaciers and determining their climatic effects will facilitate improved modeling of climatic patterns and future regional climate projections over the TP (Qian et al., 2015; Li et al., 2016; Dong et al., 2018). Provided that the parameterization schemes, such as surface albedo, soil thermal and hydraulic conductivity, and vegetation dynamics, can be revised using existing observations of cryosphere elements, land surface models will delineate land-surface processes in more detail, and then the simulations of climate models will be more realistic. This is also one of the study purposes in the Clic project initiated by the WCRP. As such, two meaningful intents will be generated. Meteorological variables output by the improved climate models can be used, on one hand, together with the TP cryosphere observations to investigate the interactions between the cryosphere and climate, and on the other hand as the atmospheric forcing for the cryosphere land-surface models, such as the glacier mass energy model (Collier et al., 2013; Mölg et al., 2014; Zhao et al., 2014; Shi et al., 2016), the snow cover model (Yuan et al., 2016), and the land surface model containing a frozen ground module (Guo and Wang, 2013; Wang et al., 2016), to realistically illustrate variations in, and response of the cryosphere to climate change (Fig. 11).

5. Summary and prospects

Studies on the cryosphere have been conducted from process description and statistical analysis to mechanism interpretation and numerical simulation (Qin and Ding, 2009). The past decades, especially those since the 1960s, have witnessed remarkable climate warming in the TP, and the cryosphere has also changed rapidly. The issues we encounter now which need further research can be divided into the following four aspects:

(1) Glacier observations in the TP have been launched for a long time. The whole variation trend is glacier retreat with apparent fluctuations against climate warming. The glacial mass balance is negative

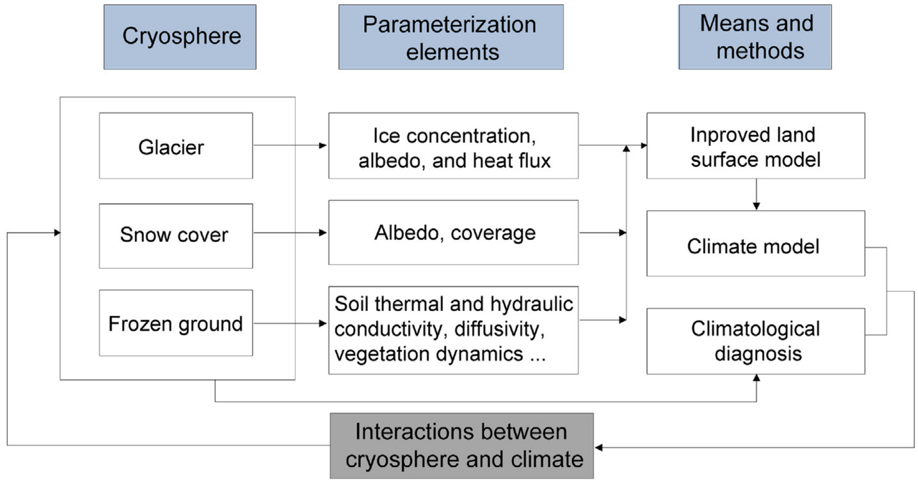

Fig. 11. Technical frame of interactions between the cryosphere and climate.

with the exceptions of the Karakoram and Kunlun Mountains and glacier melting has accelerated. The glacier retreat generally decreases from the Himalayas to the central TP and is the least in eastern Pamir. Predictions show that the area and reserve of glaciers will decrease gradually in the 21st century. The process of glacier response to climate change is exceedingly complex, depending mainly on the glacier character, type, and scale. To better use dynamic models in glacier regions, much effort is required to intensify long-term in-situ observations of glaciers at different regional scales, referring to mass balance and physical characteristics (shape, temperature, and thickness) (Li et al., 2007; Xie et al., 2009; Qin and Ding, 2009), as well as develop remote sensing techniques (Bolch et al., 2012). On the basis of employing the interaction and multi-feedback simulations from the dynamic downscaling method coupled with glacier parameters, i.e., mass balance, predicting glacier changes is a new potential and multi-scale approach to solving the issues of mountain glaciers at the regional/basin scale.

(2) Snow cover in the TP, including duration, depth, days of snow cover, and snow coverage rate, has experienced an overall decreasing trend from the 1960s to the 2010s. The general trend of snow-cover extent decreased in spring and winter and increased in summer and autumn. To fill the observational gap of snow cover, increasing the number of stations in the western TP is particularly required. Meanwhile, the features and changes of snow cover should be clearly understood by combining different kinds of data sources with inclusion of reliable satellite remote sensing data and numerical models. Of particular note is snow cover parameterizations need to be optimized in future studies (Shi et al., 2011), partially through reinforcing knowledge of LAA deposited on snow/glaciers (Qian et al., 2015; Zhang et al., 2018), aimed at addressing the remaining open questions: what role does the decreased snow cover play during the formation of rainfall patterns (Zhu and Ding, 2007) and what role do the radiation and hydrologic effects of snow cover in the TP play in the regional climate?

(3) The ground temperature of frozen ground regions in the TP has increased based on both in situ observations and simulations in past decades. The permafrost has undergone a regional degradation starting from the late 1970s. The total area of permafrost has decreased by approximately $0.46 \times 10^6\,km^2$ during 1975–2012. The permafrost layer has thinned as indicated by increased ALT. The SFD of SFG in the large portion of the TP was reduced at a rate of $0.22 \pm 0.004\,cm\,yr^{-1}$ during 1950–2009. The duration of freezing in frozen ground regions has shortened. The prediction is that frozen ground will continue its degradation under future climate warming. The assumption that ALT would increase with climate warming is noted to modify because the ALT clearly responds to summer temperature, whereas the observed warming mainly appears in spring and winter. Meanwhile, frozen ground, snow cover, and vegetation complexly interact with the ecosystem and climate (Zhang, 2005; Peng et al., 2017). For these features, it will be essential to strengthen field observation, develop large-scale observation experiments such as a long temporal series of InSAR observations and simulate the processes of frozen ground using improved numerical models that accurately describe soil water-heat physics and vegetation dynamics to dissect the frozen ground-climate interactions.

(4) Changes of the cryosphere bring far-reaching feedback to other circle layers in the climate system. Among the main elements of the cryosphere in the TP, snow cover and frozen ground greatly affect local and regional climate change. However, all the simulations only considered individual feedback effects (snow cover or frozen ground) on climate change instead of their joint effects. Therefore, how to simultaneously consider their physical processes in regional and global climate models is the key to revealing the mechanism of cryosphere influence on climate change and improving the performance of climate models. Moreover, the major feedback relationship between the cryosphere and global climate does not remain confined in physical processes (water-heat dynamics) but also intertwines with biochemistry, ecology, and geomorphologic processes, such as absorbing and releasing of greenhouse gases (CO_2, CH_4, and NO_X) (Saito et al., 2007; Shen et al., 2015). The integrated global climate (earth system) models and methods will advance the assessment of the effects of global climate change on the cryosphere (Saito et al., 2007; Qin et al., 2018).

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